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TECTONIC MOVEMENT AND DEFORMATION RELEASE IN ALASKA

by

Eduard Berg

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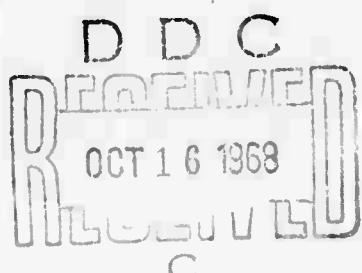
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PERSONNEL

The following persons have actively participated in the research of this program:

Larry Gedney

Susumu Kubota

Jurgen Kienle

Kenneth Hanson

Norbert Sperlich

ABSTRACT

This report covers the research effort of the seismology-volcanology section, which was supported during the past years by AFOSR grants.

The past two years have seen the steady development of seismology in Alaska. The University's Geophysical Institute has established the large telemetry network, and accurate epicenter location of even small shocks is now routine work. Crustal structure investigations in the Tanana Basin were conducted. The Fairbanks earthquakes and aftershocks in June (magnitudes 5 3/4, 6 and 5 3/4 within 20 minutes) highlighted the necessity for further studies and yielded interesting outlooks on the mechanics of pre-failure stress.

In the Katmai area--in the active volcanic belt--application of Gorshkov's idea permitted the location of magma chambers. Shear waves are screened by the reservoirs and a judicious distribution of seismic wave sources and recording sites permits the delineation of these magma chambers. A detailed gravity survey shows one of them very precisely.

In the following pages work already published or sent for publication is summarized by the abstract of the paper; research still under way, or not published is rendered in more detail.

LARGE APERTURE TELEMETERING SYSTEM FOR CENTRAL ALASKA

The telemeter system was planned in 1965 and the first stations operated in the winter of 1966-67. It was established to extend the general knowledge of seismicity and deformation release and crustal structure in Alaska. It also greatly enhances the seismic coverage of the U.S.C.&G.S. world wide epicenter location program.

The geographical coordinates of the stations were selected so as to cover the regions of highest seismicity in the Alaska Interior, the area of active volcanoes adjacent to Cook Inlet (the oil fields) and the Alaskan Peninsula (Fig. 1). These volcanoes are potentially dangerous to vital civil and military operations in the state.

The largest aperture of the system at present is 745 km (BIG-PJD). Its axis of highest resolution (perpendicular to BIG-PJD) is oriented 125° and 305°E of N (toward the Nevada test site). At the distance of the Nevada test site (some 33½°) a 50 millisecond arrival time difference of the P-wave corresponds to a lateral variation of less than 3½ km (Fig. 5). The 50 millisecond resolution between arrival times is easily obtained by cross-correlation of the first cycle of the P-arrival for events with body wave magnitudes of 4.7 and up (Fig. 6).

Experience has shown that earthquakes of body wave magnitudes of 4.0 in the 80° distance range can be detected by the routine reading of the Develocorder record, at least in favorable directions (Berg et al., July 1967). Considerable improvement in the detection capability could be derived by velocity and frequency filtering if the surveillance of a special area is required.

There is a strong suggestion that the noise level at individual stations depends largely on the gross tectonic structure. S/N variation in the network is 5:1. In addition, the sensitivity depends on the direction of the wave approach with respect to these structures. The time delayed summation of three lower and medium gain stations for the Nevada test of 23 May 1967 (SCM, BIG and TNN) shows a somewhat lower signal-to-noise ratio than the high gain PJD station alone, suggesting that the low wind areas of the geologically older formations of interior Alaska constitute ideal sites for a very high gain network.

So far, six sites are in unattended operation with completely uniform equipment. The equipment has been described by Berg et al. (July 1967). To date no equipment failure has occurred in the remote sites. Four of the six stations have never been revisited since the installation.

TABLE OF STATIONS

| <u>Name</u> | <u>Code</u> | <u>Latitude</u> | <u>Longitude</u> | <u>Altitude</u> | <u>5 cycle/sec Magnification</u> | <u>1 cycle/sec Magnification</u> |
|--------------|-------------|-----------------|------------------|-----------------|--------------------------------------|--------------------------------------|
| Tanana | TNN | 65°15.4'N | 151°54.7'W | 504 m | 0.5×10^6 | 100×10^3 |
| Big Mountain | BIG | 59°23.4'N | 155°13.0'W | 562 m | | |
| Sheep Creek | SCM | 61°50.0'N | 147°19.7'W | 1020 m | 1.0×10^6 | 115×10^3 |
| Pedro Dome | PJD | 65°02.1'N | 147°30.5'W | 740 m | 1.6×10^6 | 310×10^3 |
| Black Rapids | BLR | 63°30.1'N | 145°50.7'W | 809 m | 2.9×10^6 | 560×10^3 |
| Sparrevohn | SVW | 61°06.5'N | 155°37.1'W | | | |

Monthly Epicenter Maps

Monthly epicenter maps of earthquakes of magnitude two or over which are recorded at at least three stations have been made since February.

1967. Major areas of activity (besides Fairbanks, with its June earthquake sequence) are the southwestern part of the Alaska range, the Lake Clark Fault-Cook Inlet region and, of late, the immediate vicinity of the active volcanoes Iliamna and Redoubt (Figs. 2,3,4). Redoubt has blown several ash columns during the winter months of 1967-68. Some reached heights of 30,000 feet and were recorded on the Institute's infrasonic array systems. A computer program for depth determination is under development.

Special Data Requests from the Telemetry System

Special data have been requested most often by the U.S.C.E.G.S. in Rockville for studies of Alaska. However, occasional information has been given to the University of California at Berkeley (Dr. McEvilly) and the Texas Instrument Corporation (George Hair). The data for Texas Instruments included epicenter location and origin times on magnitude 2 to 4 shocks for study with the LASA system.

Computer Programs Available

- Epicenter determination in Alaska (no depth)
- Distance and azimuth between points (≤ 1000 km) on the earth's surface
- Calculation of Bouguer anomalies (for different densities) given station location, elevation and gravity readings
- Energy and strain release for a given time period and area (plotted curve)
- Calculation of travel times for different crustal parameters and distances.

A SEISMIC REFRACTION PROFILE AND CRUSTAL STRUCTURE
IN CENTRAL INTERIOR ALASKA

By Kenneth Hanson, Eduard Berg and Larry Gedney

ABSTRACT

An unreversed seismic refraction profile was obtained during the winter of 1966-1967 from several small chemical explosions in an open pit coal mine near Suntrana, Alaska. The profile runs from south to north across the Tanana Basin in central interior Alaska. Both P and S waves, generated by the explosions, were observed out to a distance of 217 km from the shot point. A four layer crustal model is adopted to explain the first and later prominent seismic phases. The P velocities range from 3.67 km/sec in the sedimentary surface layer to an apparent value of 8.83 km/sec in the upper mantle. It is assumed that the Moho is dipping to the south under the Alaska Range, from a depth of 32 km at Fairbanks to approximately 48 km under the shot point.

TABLE

| <u>Layer</u> | <u>v_p (km/sec)</u> | <u>v_s (km/sec)</u> | <u>Thickness (km)</u> | <u>v_p/v_s</u> |
|--------------|----------------------------------|----------------------------------|-----------------------|-----------------------------|
| 1 | 3.67 | 2.31 | 2.6 | 1.59 |
| 2 | 5.42 | 3.27 | 4.5 | 1.61 |
| 3 | 5.80 | 3.45 | 9.6 | 1.68 |
| 4 | 6.43 | 3.66 | --- | 1.80 |
| Mantle | 8.83 | 4.78 | --- | 1.85 |

THE FAIRBANKS EARTHQUAKES OF JUNE 21, 1967; AFTERSHOCK
DISTRIBUTION, FOCAL MECHANISMS, AND CRUSTAL PARAMETERS

by

Larry Gedney and Eduard Berg

ABSTRACT

A series of moderately severe earthquakes occurred in the vicinity of Fairbanks, Alaska, on the morning of June 21, 1967. During the following months, many thousands of aftershocks were recorded to outline the aftershock zone and to resolve the focal mechanism and its relationship to the regional tectonic system. No fault is visible at the surface in this area.

Foci were found to occupy a relatively small volume in the shape of an oblate cylinder tilted about 30° from the vertical. The center of the zone lay about 12 kilometers southeast of Fairbanks. Focal depths ranged from near-surface to 25 kilometers, although most were in the range 9-16 km. In the course of the investigation, it was found that the Jeffreys and Bullen velocity of 5.56 km/sec for the P-wave in the upper crustal layer is very near the true value for this area, and that the use of 1.69 for the V_p/V_s ratio gives good results in most cases.

The proposed faulting mechanism involves nearly equal components of right-lateral strike slip, and normal faulting with northeast side downthrown on a system of sub-parallel faults striking N40°W. The fault surface appears to be curved--dipping from near vertical close to the surface to less steep northeast dips at greater depths. The relationship of this fault system with the grosser aspects of regional tectonism is not clear.

EVIDENCE FOR MAGMA IN THE KATMAI VOLCANIC RANGE

S. Kubota and E. Berg

ABSTRACT

Several independent observations during the summer of 1965 suggest the presence of magma in the volcanic range of Katmai. A high value of 0.3 for Poisson's ratio and the screening of predominantly the vertical component of the elastic shear waves have been observed. Narrow negative Bouguer anomalies possibly indicate the presence of low density material at shallow depth. The location of magma reservoirs has been attempted using the calculated wave path and the screening of the mainly vertically polarized shear wave.

Of the possible ten chambers thus located, the ones of shallow depth (to 20 km) correspond to the location of active volcanoes. The ones between the 20 km level to the upper mantle seem to spread over a rather wide area and are not clearly related to the geographical position of a particular volcano.

Theoretical considerations on the propagation of elastic waves substantiate the observed absence of vertically polarized shear waves.

PRELIMINARY DETERMINATION OF CRUSTAL STRUCTURE
IN THE KATMAI NATIONAL MONUMENT, ALASKA

by

Eduard Berg, Susumu Kubota* and Jurgen Kienle

ABSTRACT

Seismic and gravity observations were carried out in the active volcanic area of Katmai in the summer of 1965. A determination of hypocenters has been attempted using S and P-arrivals at a station located at Kodiak and two stations located in the Monument. However, in most cases, deviations of travel times from the J-B tables were rather large. Therefore hypocenters are not well located. A method based on P and S-wave arrivals yields a Poisson's ratio of 0.3 for the upper part of the mantle under Katmai. This higher value is probably due to the magma formation. The average depth to the Moho from seismic data in the same area is 38 km and 32 km under Kodiak. Using Woppard's relation between Bouguer anomaly and depth to the Moho, a small mountain root under the volcanoes with a depth of 34 km was found dipping gently up to 31 km on the NW side. The active volcanic cones are located along an uplift block. This block is associated with a 35 mgal Bouguer anomaly. The Bouguer anomaly contour map for the Alaska Peninsula is given and an interpretation attempted.

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FORESHOCKS, STRESS LEVELS

Introduction

In recent years an increasing number of laboratory fracture experiments on small rock samples have been performed. The results show that the Gutenberg-Richter relation

$$\log N = a + b (S - M)$$

holds for a large variety of rock types over several orders of magnitude.

In this expression the constants "a" and "b" relate to the level of "seismicity" and to the physical properties and stress state of the rock. N is the number of events with magnitudes larger than or equal to magnitude M.

Mogi (see several refs.) showed that the slope "b" obtained for microfracturing in rock samples is similar to the familiar values found in earthquake seismology. These slopes range from 0.7 to 1.1 for "normal" seismicity and aftershock series to between 2.0 and 3.0 for very shallow volcanic earthquakes. Only recently have very small values of 0.3 - 0.5 been found. These results suggest that the theory describing the failure mechanism in both small rock samples and earthquake zones must contain similar statistical elements. Mogi also found the same general behavior in the microfracturing of his samples as occur with natural earthquakes. One of the following "typical" sequences occurred: 1) no "foreshocks", large main shock and many aftershocks, 2) some foreshocks, large main shock and some aftershocks, 3) earthquake "swarms", with many small shocks increasing in number with time and then subsiding. He associated the above cases, respectively, with fracturing in material exhibiting:

- (1) homogeneity
- (2) partial inhomogeneity allowing stress concentration at crack tips

(3) very high heterogeneity where no large stress build up is possible.

The strain release through fracturing in the second case increased rapidly prior to the large "main shock" for a variety of materials. The same behavior has been observed for the tectonic block between 179°E and 165°W in the Aleutian Chain prior to a large earthquake in 1957, and in mainland Alaska in 1912-1913, and in the early thirties (Berg, May, 1966).

Scholz (February, 1968) recently investigated microfracturing in laboratory samples, varying the applied stress from zero to the breaking point. His findings substantiate Mogi's in that the slope "b" of the Gutenberg-Richter equation was found to vary similarly for different rock samples. The most striking conclusion was that the slope directly reflects the stress state of the sample. At low stress levels he attributes microfracture activity to frictional sliding on pre-existing cracks and to the crushing of pores, a situation similar to the occurrence of shallow volcanic earthquakes with high "b" values from 2 to 3. At 30 to 50% of the fracture strength, few events were detected. Above 50%, microfracture rate increases continuously until rupture occurs, and the "b" values decrease at the same time from about 1 to smaller values with increasing stress. The same general decrease of "b" was found in samples which were confined under pressures up to 5 k bars, but due to frictional sliding along pre-existing cracks. Scholz uses a statistical distribution of stress, given the mean stress, to derive the Gutenberg-Richter relation. One interesting result is that the numerical value of "b" for a single component mean stress is confined within certain limits. Using an empirical derived value for a constant (in his eq. 14), "b" is limited to values between 1 and 0, in good agreement with non-volcanic earthquake seismicity.

In consideration of the above, over 60 foreshocks of the June 21, 1967 Fairbanks earthquake were analyzed and compared with the results found for other foreshock series in studies by Suyehiro (1964, 1966) and Papazachos et al. (1967). Suyehiro computed "b" values for the foreshocks of a relatively small magnitude 3.3 quake in Japan, and of the magnitude 8.5 Chilean quake in May 1960. In both cases these values were about half those obtained from the aftershock series. The observed precursors began 4 and 33 hours, respectively, prior to the main shocks. Only 23 foreshocks for the Japanese earthquake, and 31 for the Chilean, were reported. The data presented on the foreshocks of the Kephallenia Earthquake ($M = 7 \frac{1}{4}$) by Papazachos have a relative large spread. On the basis of 50 foreshocks, the authors obtain a value for "b" of 0.61, as compared with 0.85 for the aftershocks. Data for the above examples are summarized in Fig. 10. Ryall et al. (1968) pointed out the limited accuracy of "b" expected from the relatively small sample size. The mean square error determined for "b" diminishes with increase in sample size. For example, Ryall found that a mean square error of ± 0.24 results from samples of 50 shocks, while 0.07 is the value if 99 shocks are used. Two samples of 200 shocks each were then investigated, and the values of "b" differed by only 0.02. In this case the different sized samples were all obtained from the same body of data. The slope "b" was 0.77. For smaller slopes one would expect the percentage of error to be roughly the same.

Three earthquakes of magnitude $5 \frac{1}{4}$ to 6.0 occurred on June 21, 1967 near Fairbanks. They were followed by a large number of felt and recorded aftershocks (Berg et al., 1967). A high sensitivity station near the epicentral area (Pedro Dome, PJD) had been installed in mid-January of the same year.

Aftershock recordings at that station exhibited $T_s - T_p$ times of 2.5 to 4.5 sec. Very small shocks in this $T_s - T_p$ time range had been recorded since the installation of the station. They are herein considered to be foreshocks of the June 21 earthquakes. The magnitude of the Fairbanks main shock ($M = 6.0$) is midway in the range of magnitudes considered in the foreshock studies by Suyehiro and Papazachos et al. In the case of the Fairbanks earthquakes, the data should be considered reliable since both foreshock and aftershocks were measured from the same seismometer and recording equipment without change in filter or magnification setting.

Location of Fairbanks Fore and Aftershocks and Data Selection

Aftershock recordings had been carried out at several temporary sites (Berg et al., 1967; Gedney and Berg, (in press)). The aftershock zone is roughly 10×15 km and all shocks occurred in the crust (Fig. 7). Recorded $T_s - T_p$ times range from 2.5 to 4.5 sec at PJD. Both foreshocks and aftershocks were measured on PJD Helicorder records. The aftershocks investigated included some 200 during the period 2013, 3 July to 1216, 4 July, and 190 during the period 2332, 2 December to 2312, 8 December.

Since recording with only one station did not permit epicenter location, all foreshocks recorded at PJD with $T_s - T_p \leq 10$ sec were investigated. There is a preponderance of events with $T_s - T_p$ time interval around 3 and 4 seconds. None of the later recordings with the temporary network showed any earthquakes with $T_s - T_p$ times in the 2.5-4.5 sec range at PJD which were not aftershocks from the June 21 quakes. We strongly feel that the shocks occurring since the installation of PJD, and until the main quake in that $T_s - T_p$ interval, must be considered as foreshocks. During the first two weeks after the installation

of PJD in January, 1967, we experimented with gain and filter settings and events recorded during this time were not considered. Some recording gaps exist during the following months due to high microseism levels or use of the Helicorder as a direct monitor for another station. The total number of foreshocks actually occurring during this period is therefore probably larger than that used in the investigation.

Two criteria were used in selecting the aftershocks to determine "b": "readability" of records and time lapse from main shock. First, a record without too many overlapping quakes and with low microseismic level was selected for the first group of some 200 earthquakes during 3-4 July. Second, in order to find if a gross variation of "b" with time exists, the early December record was chosen. Arrival times were read to the nearest minute, $T_s - T_p$ times to the nearest tenth second, and half-amplitudes to the nearest tenth of a millimeter.

Frequency of Occurrence - Magnitude Relation

The variation in distance from PJD to the maximum limits of the epicentral area is not great (about 15 km). Therefore it was considered of negligible influence on the amplitude of the recordings. Instead of converting amplitude to magnitude, the log of the amplitude versus the log of the cumulative number of events was plotted. The log of the S-wave amplitude was broken into increments of 0.1, corresponding to magnitude increments of 0.1 in conventional studies. The scaling intervals used in plotting progressed in the following fashion: $\log_{10} A = 0.4$ if the amplitude ranged from 2.5 to 3.1 mm; $\log_{10} A = 0.5$ for A between 3.2 and 3.9 mm, etc. Since only a short magnitude interval was considered, no corrections have been applied for change in frequency.

The data cover about 1.5 magnitude range but fell short at both ends. The slope "b" could therefore be graphically obtained for roughly 10 data points over a one magnitude interval.

The influence of the sample size was discussed in the introduction. From this, the 200 quakes which occurred early in July were judged sufficient to determine "b" of the aftershock series. However, to find any gross variation with time a second sample was included from shocks in early December. There was no detectable change in slope between aftershocks of these two time periods. The slope found from the aftershocks was 0.90 ± 0.04 , whereas that found from the foreshocks is 0.40 ± 0.04 . The results are presented in Fig. 9.

Obviously the difference in the rate of occurrence with decreasing magnitude of foreshocks and aftershocks is quite marked and falls well outside of any variation which might be considered due to the sample size: From Ryall's et al. (1968) discussion, a mean error of the slope for a sample size of 66 would be ± 0.15 if a sufficient number of samples of that size were available and the slope is in the vicinity of 0.8. However as the absolute value of the slope "b" decreases to roughly half that value, or 0.4, the mean square error should be closer to ± 0.08 , i.e., the same percentage error.

Discussion

1. Duration of Foreshocks

The Fairbanks earthquake's magnitude ($M=6.0$) ranges half way between those of Suyehiro's investigation. There is no definite beginning of foreshocks in this case compared to the relatively short time intervals of 4 and 33 hours found in his investigation. In the Greek case presented by Papazachos, foreshocks started 3 days prior to the main shock. Foreshocks of the June 21

Fairbanks quake occurred at least since the installation of the PJD station in mid-January of the same year, some 5 months prior.

Therefore, two time scales are involved in the forewarning pattern of strain release: first, one with a duration of hours to a few days; second, one covering on the order of half a year (or possibly longer). Short term phenomena preceding ultimate failure include anomalous tilts (Karmaleeva, Nishimura) and change in the magnetic field intensity in the vicinity of the fault zone. Triggering of earthquakes by ocean tidal loads (Berg, 1966; Matuzawa, 1964) and solid earth tides (Ryall et al., 1968) has been postulated.

Long range phenomena preceding failure by half a year or so are tilts toward or away from the epicentral region and rapid strain release increase (Berg, May 1966). Slow deformation of geodetic networks has been noted. The increasing rate of strain release through earthquakes clearly points out that larger parts of the tectonic block are reaching failure stress or that the average stress in the block is increasing.

2. Slope and Stress

The difference of the slope "b" between fore and aftershocks in the case of the Fairbanks earthquakes is significant and outside possible variations due to sample size. This is the first time that such a large number of foreshocks has been available. The number of aftershocks in each of two samples was chosen sufficiently high to avoid any significant error in slope. The value for aftershocks found (0.90) compares to a world average of 0.90 ± 0.02 for shallow shocks (Gutenberg and Richter, 1954) and 1.1 ± 0.02 given by the same authors for Alaska. In a more recent investigation covering 60 years, a value of 1.00 was found for Alaska, including the Aleutians (Berg,

1965). This high value of the slope "b" of aftershocks contrasts strikingly with the value of 0.40 ± 0.04 for the foreshocks (Fig. 9). Figure (10) shows a clear areal dependence of the value for "b". Mogi (see several references) has pointed out the striking similarities between the fracture statistics obtained in the laboratory from rock specimens and the earthquake statistics. Berg (May 66) found that some of these results apply in the earthquake belt of the Aleutians and Alaska. Scholz has further contributed to the understanding of microfractures in laboratory experiments through explanations of the contribution of physical parameters such as stress and confining pressure as parameters for the frequency-magnitude relation. He developed a statistical model that fits the observed Gutenberg-Richter relation for the microfractures.

It now appears that there is a sufficient body of evidence to postulate a definite link between the slope of the frequency-magnitude relationship of foreshocks and aftershocks.

If Scholz's findings for the dependence of microfracture occurrence on stress and confining pressure conditions is applicable to the much larger earthquake zone, the following conclusion can be drawn from the investigation of the Fairbanks earthquakes: Foreshocks contribute only negligibly to the stress drop in the stressed volume. The main stress-drop occurs during the main shock, and stress conditions, though considerably lessened, remain relatively unaffected by an aftershock sequence (at least--in the case of the Fairbanks earthquakes--during the short period from July to early December, 1967).

There is also indication that the percentage of fracture stress difference sustained in a stressed earthquake region may decrease with increasing volume. The slope "b" of the frequency-magnitude relationship decreases with

increasing percentage of fracture stress difference with all materials tested by Scholz. His Fig. 4 (Fig. 11) shows this decrease for pressures (up to 5 kb) similar to those encountered in the crust. If the results of the laboratory measurements (on Westerly granite) can be applied to earthquakes in the crust, a further conclusion is that the mean stress difference in the case of the Japanese and Fairbanks foreshocks reached roughly the 90% and 80% fracture-stress difference level prior to the main shock. In the absence of tilt and/or strain or geodetic measurements, the slope "b" of the foreshocks also may be an indication of the expected size of the main shock.

In an earthquake situation foreshock sequences with small values of "b" preceded small main shocks. The lab experiment indicates a high percentage of fracture stress difference for such a small "b". Foreshock sequences with large "b" values occur prior to larger main shocks. Aftershock volume increases with the magnitude of the main shock. The preceding therefore seems to be an indication that the percentage of fracture stress difference which is sustained in a stressed earthquake volume decreases with increasing volume. Or, stated in a different manner: small crustal volumes may be able to sustain higher stress levels than larger ones. If foreshocks occur at all, larger areas must already have built in failure zones (cracks) and the relative stress concentration at the ends of such zones (cracks) is larger for extended ones than for small ones. Therefore, if a foreshock sequence exhibits a high "b" value one might expect that the dimensions of the stressed volume are relatively large and that failure in a large zone or volume is possible. Conversely, a small value for "b" in a foreshock series would indicate that a relatively high percentage of fracture stress difference level exists in the stressed volume, which is only possible for relatively small zones or volumes.

Additional indication of the extent of the stressed volume may be obtained from the foreshock hypocenter distribution. The preceding conclusions are tentative, however, and need more data to be fully supported.

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Fig. 1

Seismograph stations.

GEOPHYSICAL INSTITUTE, UNIVERSITY OF ALASKA

● TELEMETER STATION
○ OTHER STATION



0 100 200 300 400 500 Miles
0 200 300 400 Kilometers

FOR SALE BY U. S. GEOLOGICAL SURVEY, FEDERAL CENTER, DENVER, COLORADO OR WASHINGTON 25, D. C.

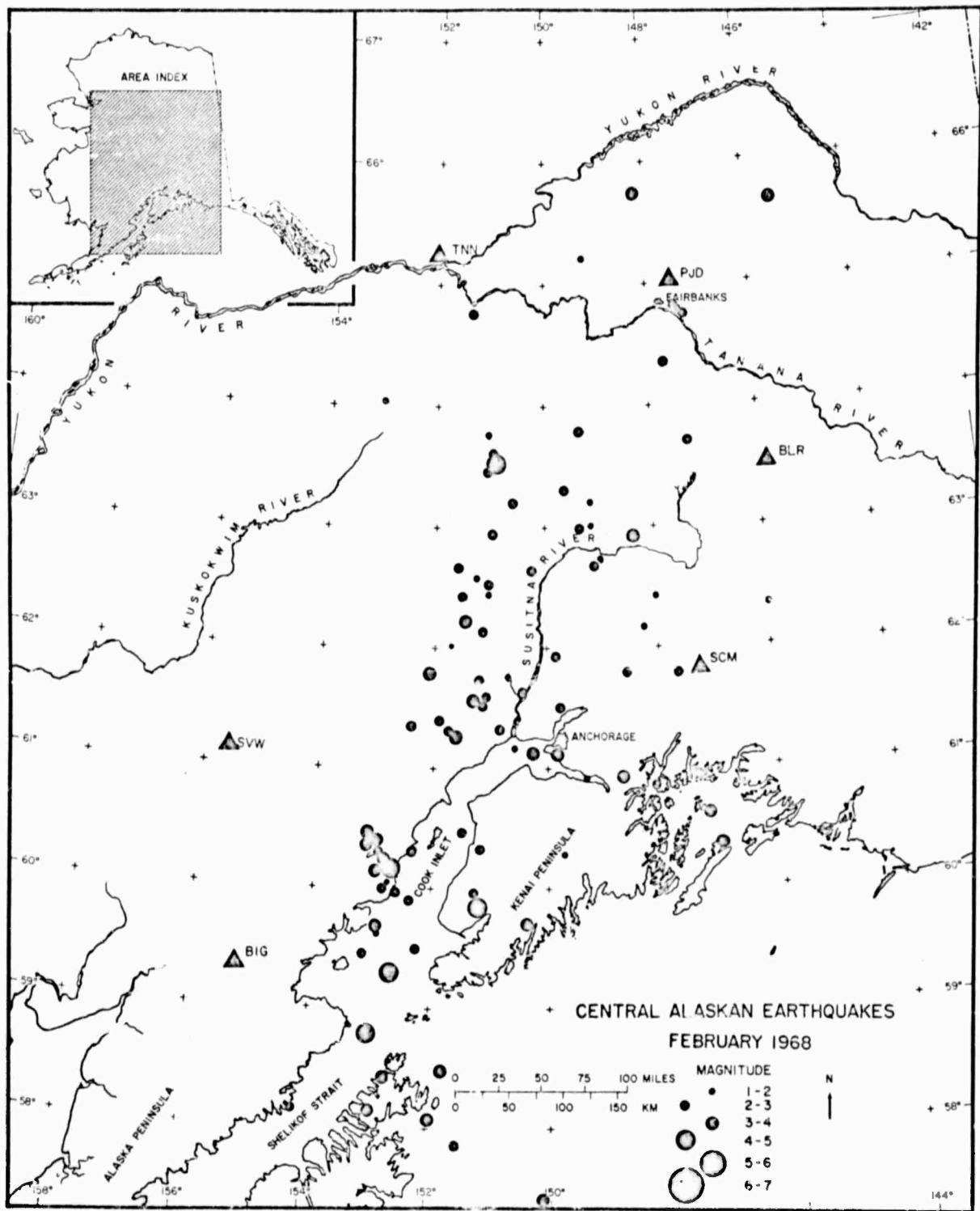


Fig. 2 Monthly epicenter maps.

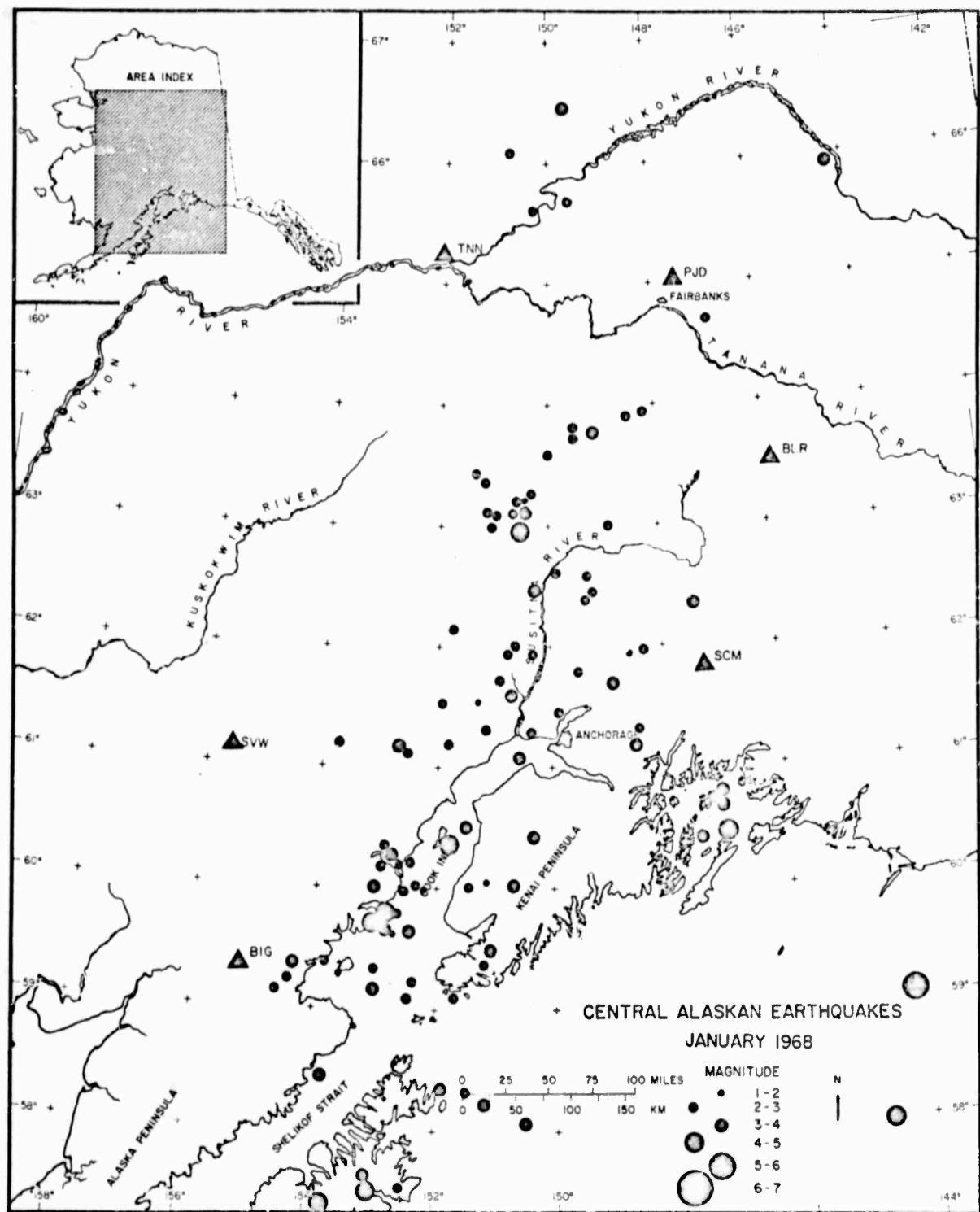


Fig. 3 Monthly epicenter maps.

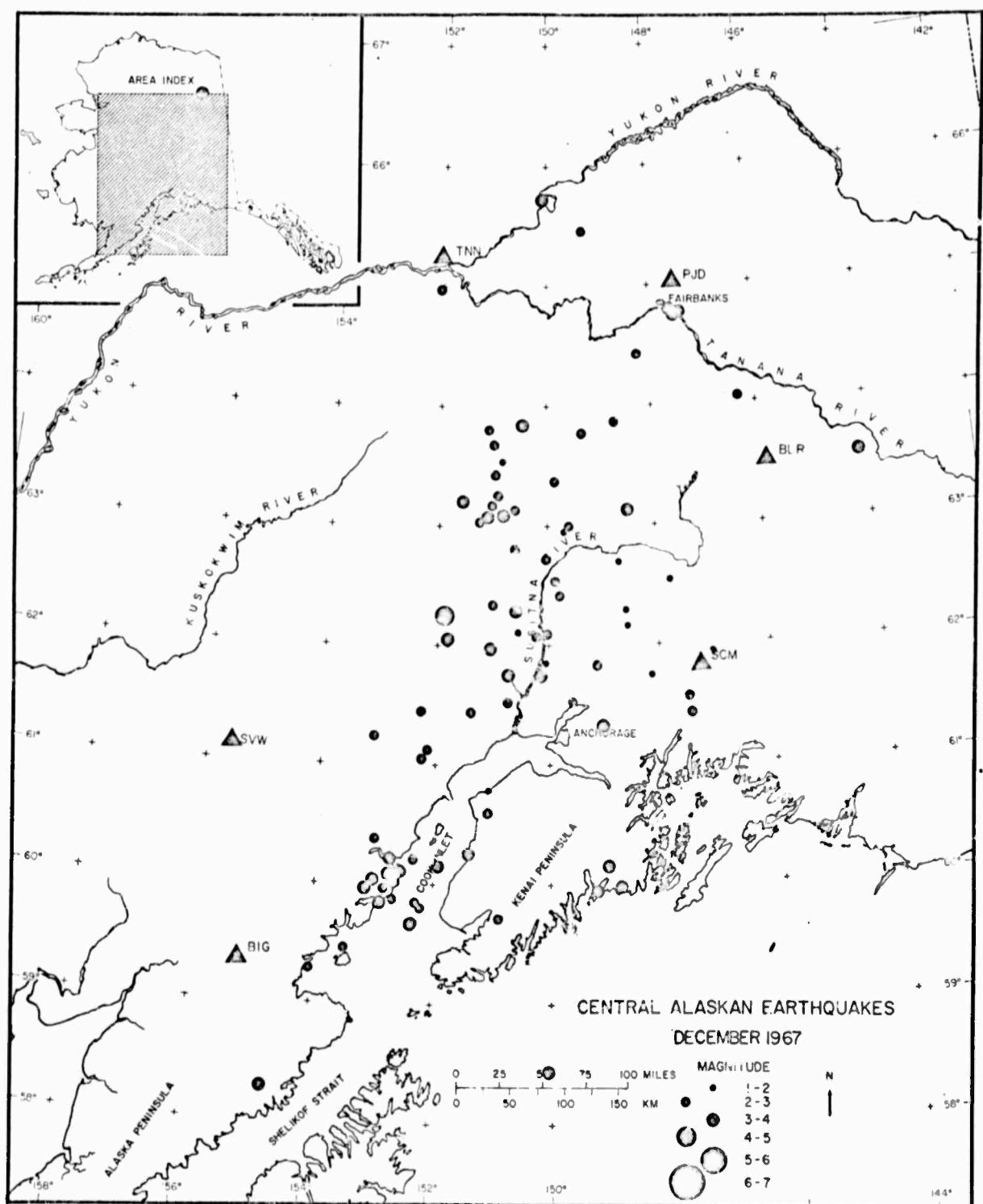


Fig. 4 Monthly epicenter maps.

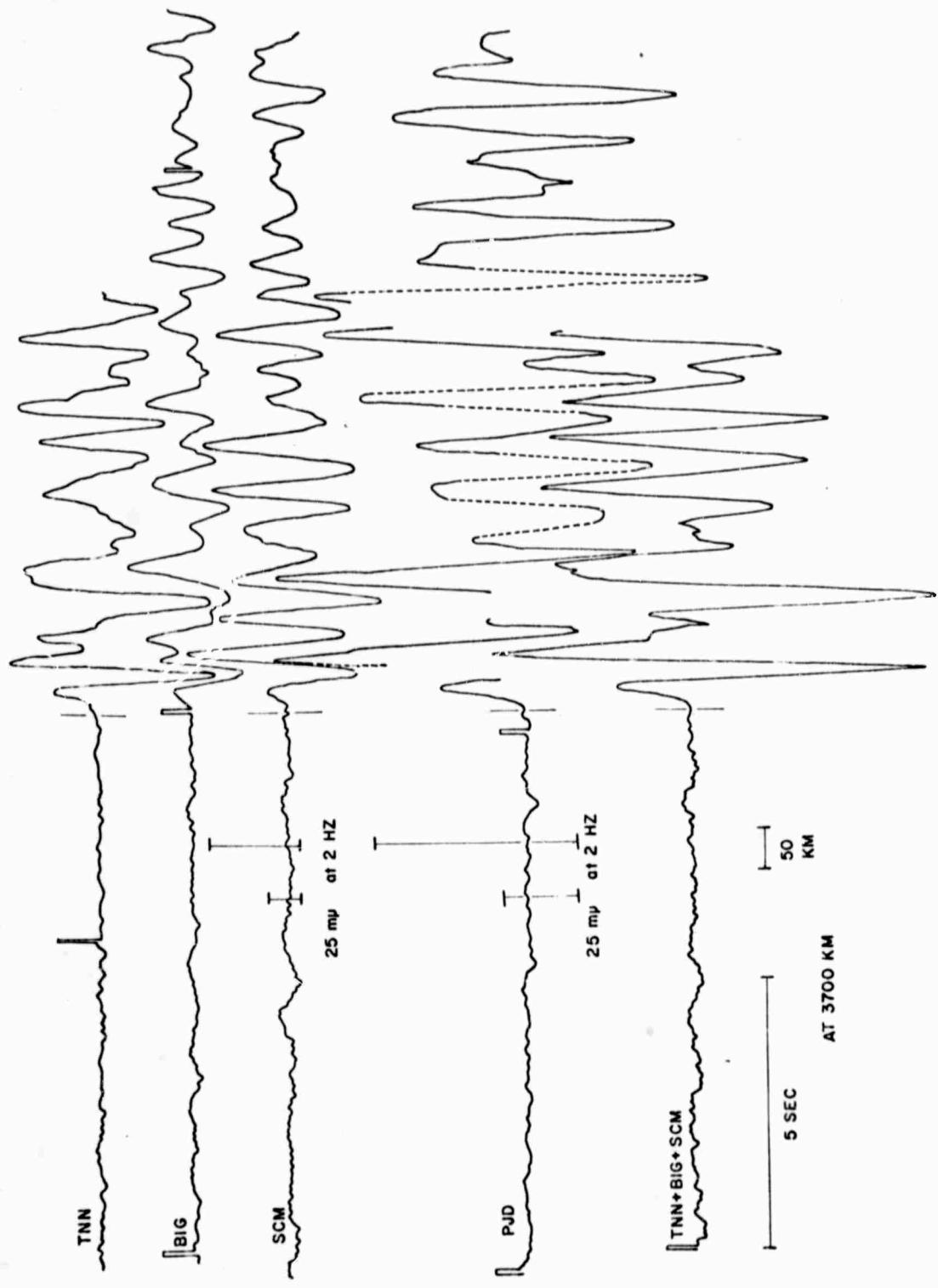


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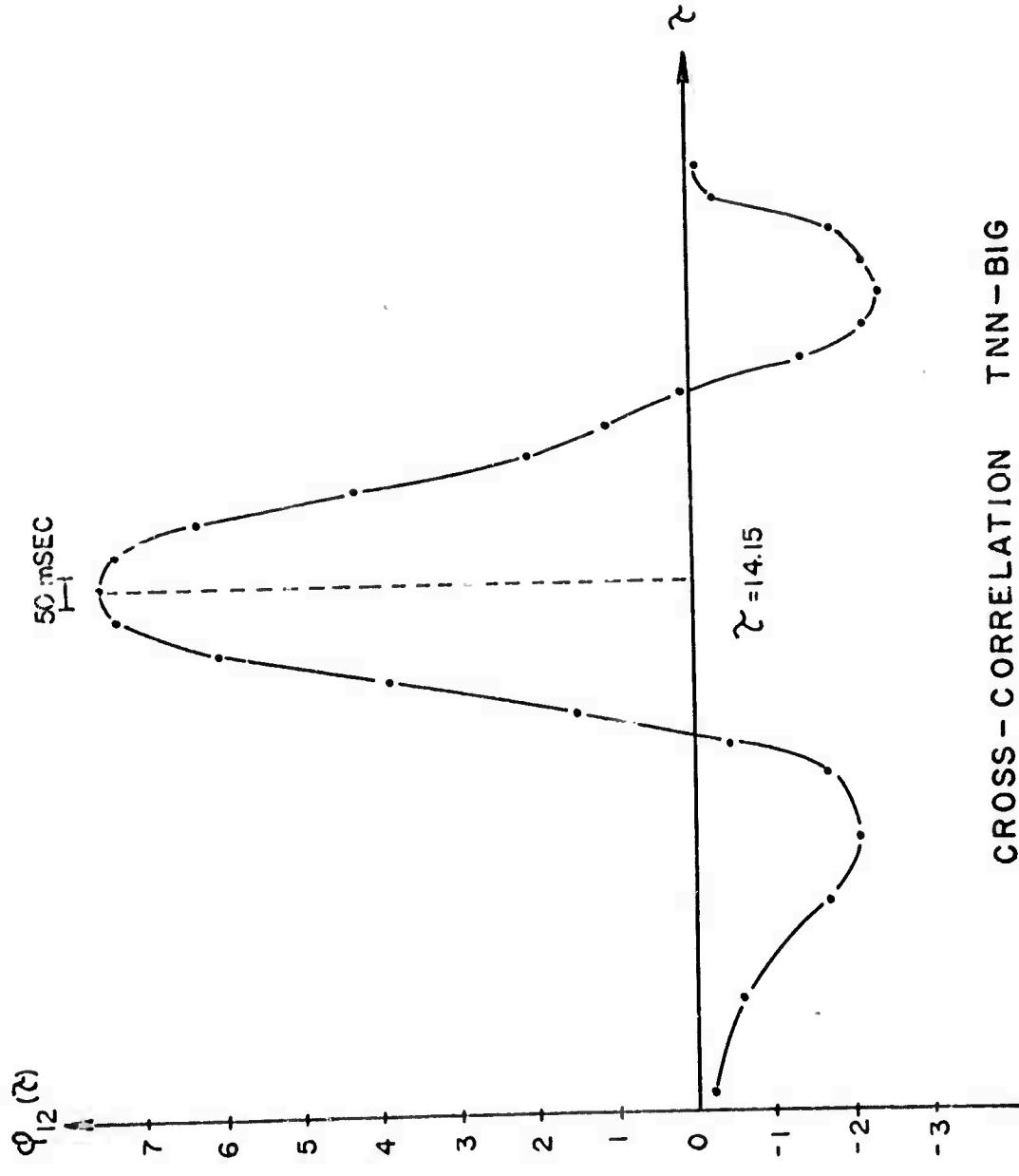


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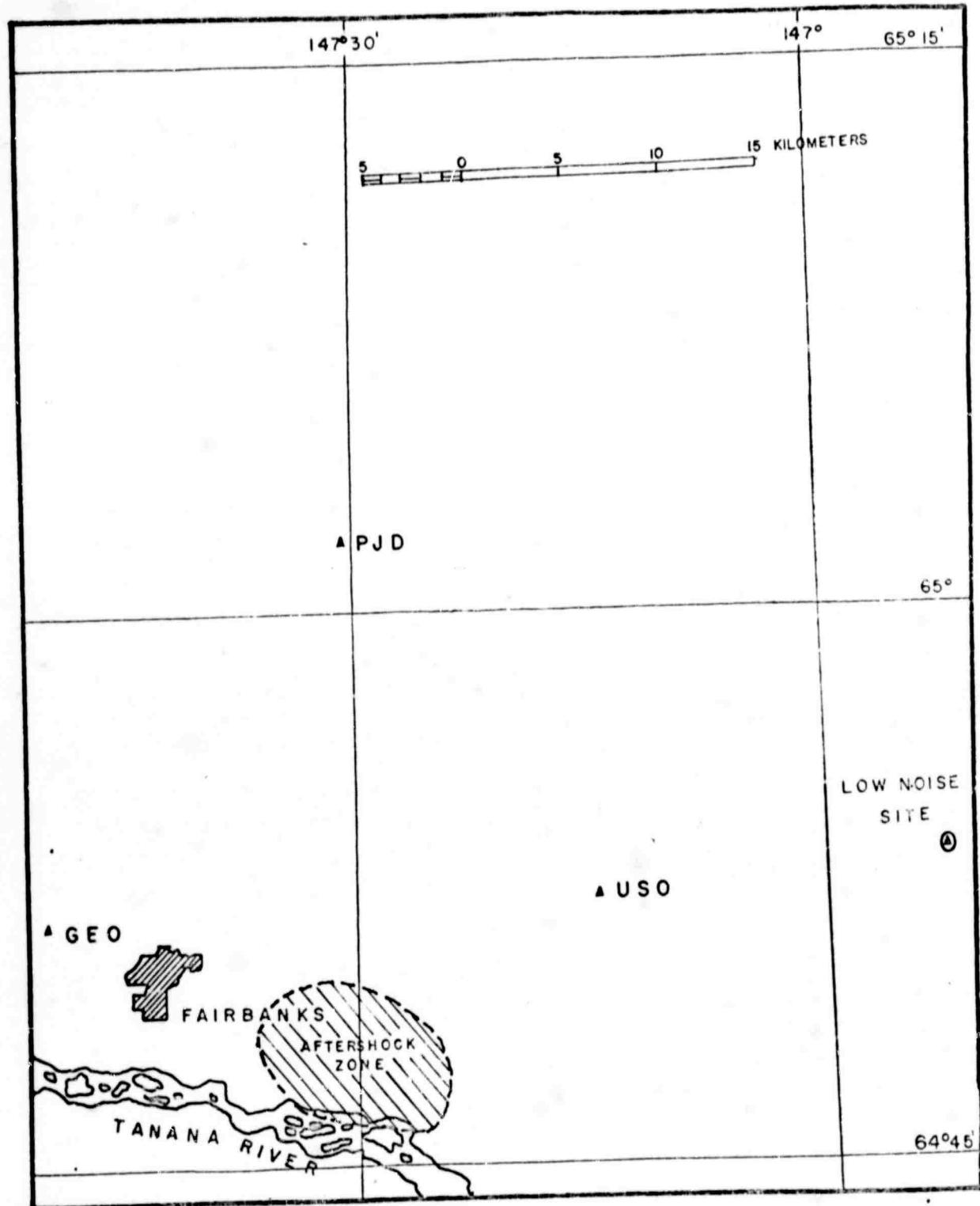


Fig. 7 Fairbanks earthquake aftershock zone.

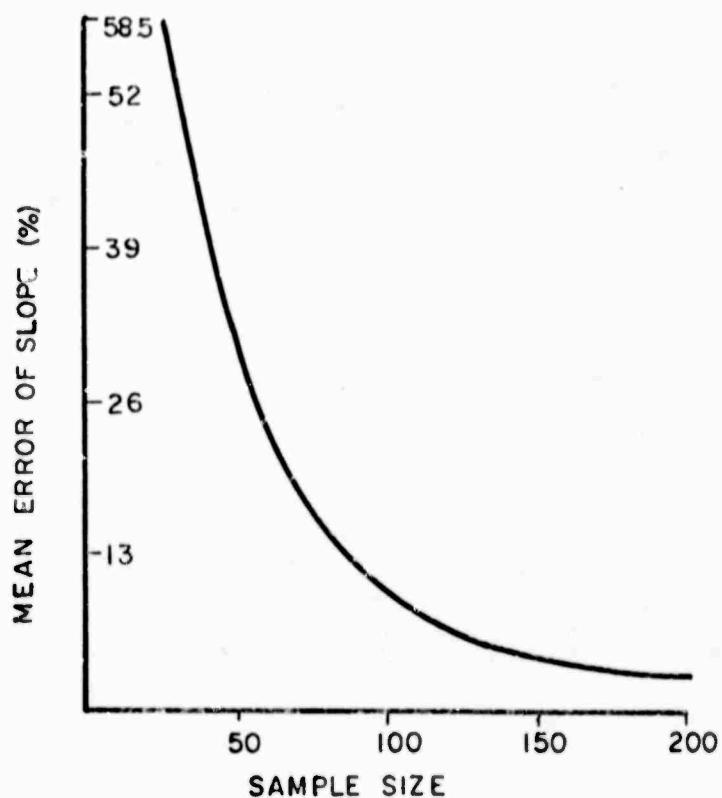


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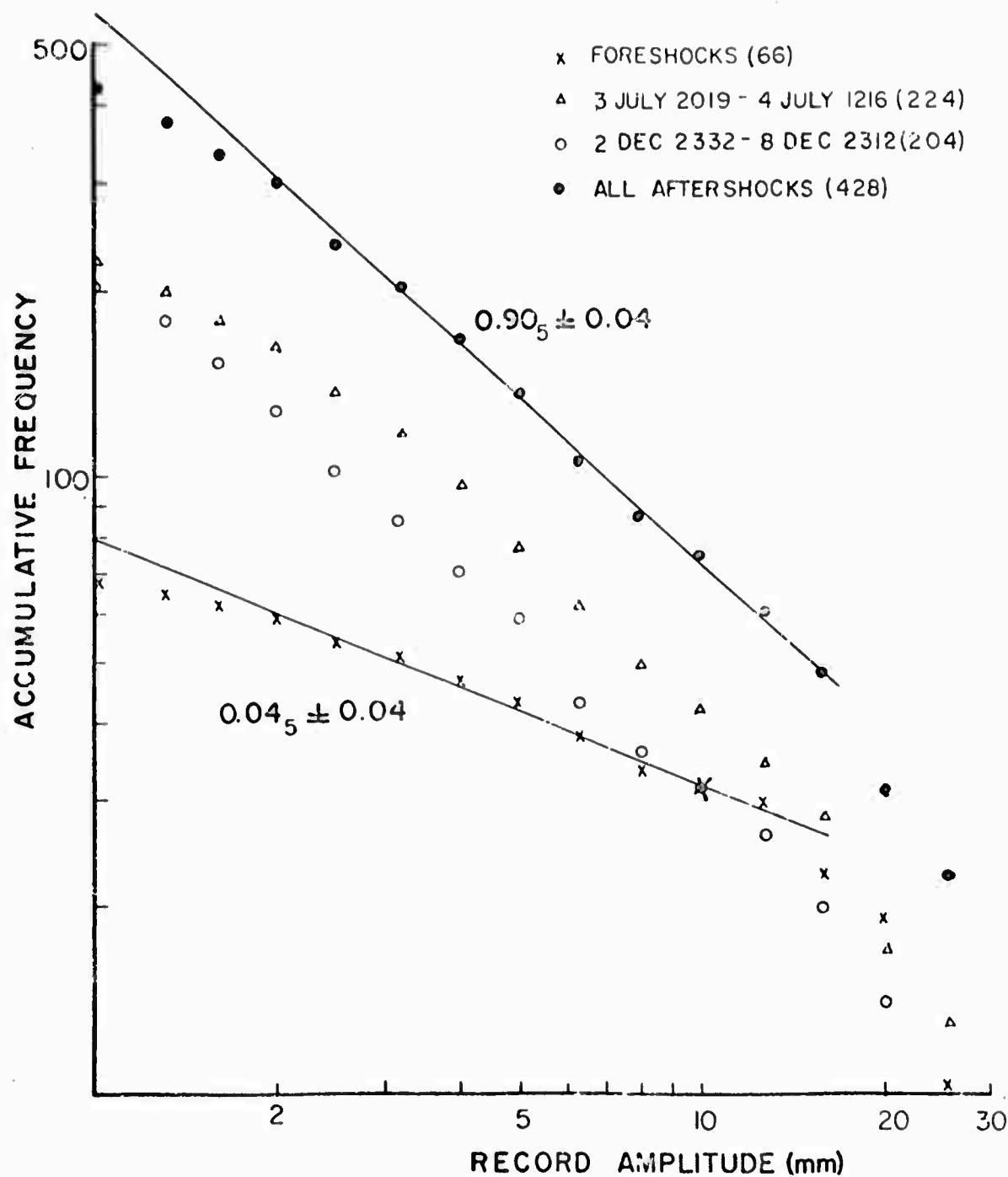


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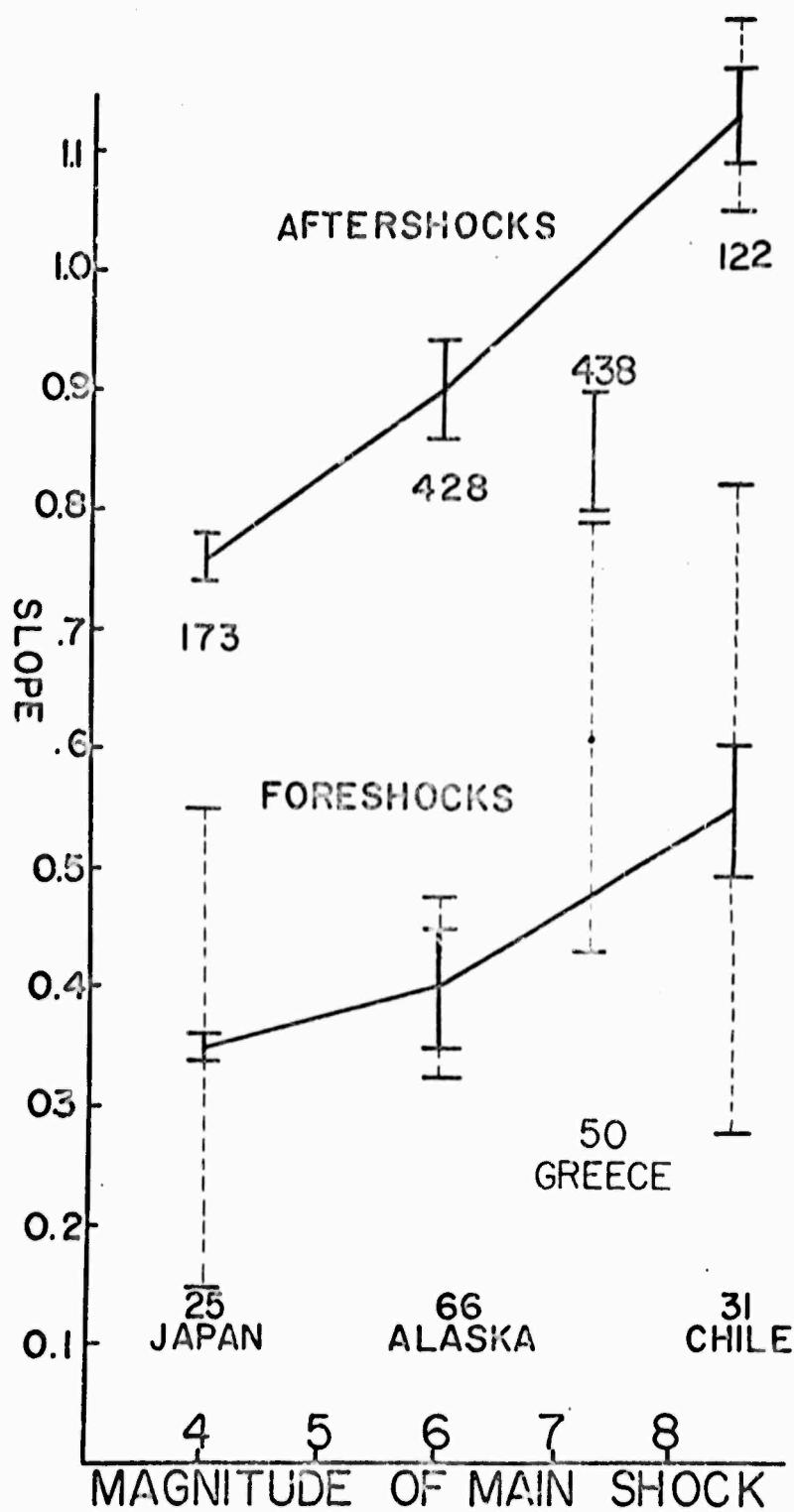


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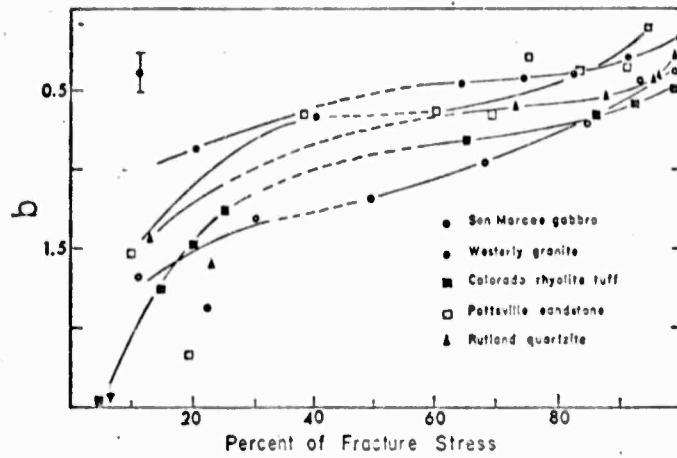


FIG. 3. b as a function of normalized stress for five rocks in uniaxial compression. The dashed part of the curves are in the region where few events were detected.

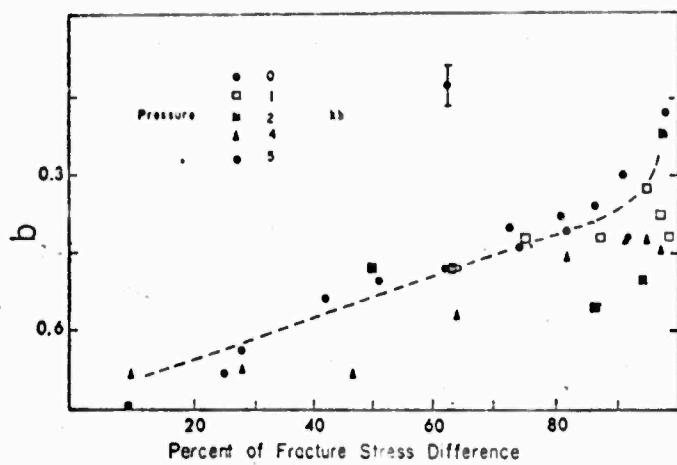


FIG. 4. b versus normalized stress difference for Westerly granite at five different confining pressures.

Fig. 11 Slope "b" as function of percent fracture stress difference (from Scholz 1968).

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| 13. ABSTRACT This report covers the research effort of the seismology-volcanology section, which was supported during the past years by AFOSR grants. | | |
| The past two years have seen the steady development of seismology in Alaska. The University's Geophysical Institute has established the large telemetry network, and accurate epicenter location of even small shocks is now routine work. Crustal structure determination in the Tanana Basin were attempted. The Fairbanks earthquakes and aftershocks in June (magnitude 5 3/4, 6 and 5 3/4 within 20 minutes) highlighted the necessity for further studies and yielded interesting outlooks on the mechanics of pre-failure stress. | | |
| In the Katmai area--in the active volcanic belt--application of Gorshkov's idea permitted the determination of the magma chambers. Shear waves are screened by the reservoirs. A judicious distribution of seismic wave sources and recording sites permits the delineation of these magma chambers. A detailed gravity survey shows one of them very precisely. | | |
| In the following pages work already published or sent for publication is summarized by the abstract of the paper; research still under way, or not published is rendered in more detail. | | |

| 14. | KEY WORDS | LINK A | | LINK B | | LINK C | |
|-----|--|--------|----|--------|----|--------|----|
| | | ROLE | WT | ROLE | WT | ROLE | WT |
| | Alaska 745 km Aperture seismic telemeter network Resolution Seismicity Fairbanks earthquake Crustal structure Magma detection gravity Foreshocks, stress, mainshocks | | | | | | |